Stress Magnitude Estimates From Earthquakes in Oceanic Plate Interiors

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We propose a method to estimate stress magnitudes in oceanic plate interiors from focal depths and focal mechanisms. Using a depth-dependent rheology, we show it is possible to estimate the differential stress \((\sigma_1 - \sigma_2)\), averaged over some reference lithospheric thickness. The resolving power of the method is investigated by evaluating the effect of uncertainties in parameters that are involved in the analysis. We apply the method to the Central Indian Ocean, where intraplate seismicity is high. From well-studied earthquakes we estimate differential stresses of the order of hundreds of megapascals. This result is consistent with the high level of stress that was found from numerical model calculations by Cloetingh and Wortel (1985, 1986). From the few intraplate events in the Pacific plate, we also estimate differential stresses in this area.

INTRODUCTION

Earthquakes relax stresses within the lithosphere, and therefore information on the stress distribution at the hypocenter is contained in seismograms. The relevance of intraplate stresses in the context of plate dynamics is an important motivation to study earthquakes in the interior parts of the plates. Directions of principal stresses are generally derived from focal mechanisms, either directly from \(P\) and \(T\) axes or via the criterion of Raleigh et al. [1972] if one of the nodal planes can be identified as the fault plane. The general consistency of stress trajectories obtained from focal mechanism data and their good agreement with stress orientations derived from various other indicators show that such directions have a regional significance [Zoback et al., 1989].

However, a more complete specification of the stress field also requires magnitude information.

We propose a method to estimate differential stress magnitudes in oceanic plate interiors. By relating focal depth and focal mechanism information to a depth-dependent rheology, it will be shown that we can estimate the differential stress level in the lithosphere where an earthquake occurs. Hence, starting from high-quality seismological data and a model for the rheology of the lithosphere (which involves several assumptions and simplifications), we derive estimates for the stress level in the lithosphere. Composition and temperature structure within the lithosphere need to be known well enough to obtain a good approximation of the strength distribution with depth. Compared with continental lithosphere, oceanic lithosphere is therefore more suited for application of the method, although no principal objections exist to using it in continental areas. In this paper we focus on estimating differential stress magnitudes in oceanic plate interiors. Well-resolved earthquake depths are a prerequisite for the method, so that intraplate earthquakes that have been the subject of depth relocation studies provide the best data for estimating stress magnitudes in oceanic plates.

The high seismicity in the Central Indian Ocean [e.g., Wiens and Stein, 1983, 1984; Bergman and Solomon, 1984, 1985] makes this area very well suited for studying the relation between seismicity and differential stress magnitudes. In a force modeling study of the Indian plate, Cloetingh and Wortel [1985, 1986] calculated stresses in the Central Indian Ocean that are very high (several hundreds of megapascals), in comparison with studies by Richardson et al. [1979] and Richardson [1987, 1989].

In agreement with Cloetingh and Wortel's results, Zuber [1987] and McAdoo and Sandwell [1985] found that hundreds of megapascals stresses are required to explain the gravity highs in the Central Indian Ocean that are attributed to basement undulations as a result of lithospheric buckling. We will test whether Cloetingh and Wortel's stress results are approximately an order of magnitude too high, as suggested by Richardson [1987, 1989]. Therefore, at stages in the analysis when assumptions need to be made, we will adopt those assumptions which (1) are considered to be realistic and (2) give low-end differential stress estimates. Finally, we will estimate differential stress magnitudes for the Pacific Ocean from its seismicity, which is significantly lower than in the Central Indian Ocean.

RHEOLOGY

In relating the depth of an oceanic intraplate earthquake to the stress field in the lithosphere, the rheology model adopted strongly affects the inferred differential stress values. In this section we briefly describe the rheology model adopted in this paper. Brittle deformation in oceanic lithosphere is expected to occur at shallow depth. We assume that the strength \((\sigma_1 - \sigma_3)\) in the brittle regime is described by Byerlee's [1978] law. At deeper levels deformation occurs by temperature-activated ductile flow, controlled by creep of olivine [Goetze and Evans, 1979; Kirby and Kronenberg, 1987].
Brittle Strength

The strength of brittle rock is to a first approximation insensitive to temperature and is mainly pressure-controlled. The stress required for initiation of slip on pre-cut rock faces was found by Byerlee [1978] to be insensitive to rock type. A linear frictional law provides a useful approximation of brittle strength in the lithosphere, particularly at effective pressures above approximately 100 MPa, where the initial surface roughness has little or no effect on friction [Byerlee, 1978].

An estimate of the pore fluid pressure at the fault surface is required to calculate the brittle strength with Byerlee's law. Fluid pressures may range from zero to superlithostatic and are not known in general. Lacking detailed information on pore fluid pressure distributions in the seismic source regions, we assume that the pore fluid pressure is hydrostatic. As the mantle source region for the growth of oceanic lithosphere is very likely to be extremely depleted in volatile elements [Anderson, 1989], very little water is available for building substantial fluid pressures in sub-crustal oceanic lithosphere [Dixon et al., 1988]. Ophiolitic rocks show evidence of hot water percolation in the crust, where an open system of pores is likely to exist. Therefore, fluid pressures probably are close to hydrostatic in the upper 5-10 km and subhydrostatic in the rest of the oceanic lithosphere. Our assumption of hydrostatic pore fluid pressure therefore yields a lower limit to the brittle strength and is consistent with our approach of getting low-end stress estimates, within the limits of (what we consider to be) realistic assumptions.

The brittle strength \((\sigma_1 - \sigma_3)\) depends on which of the principal stresses is vertical. We adopt the convention that compression is positive and that \(\sigma_2 > \sigma_1 > \sigma_3\). If \(\sigma_3\) is vertical, the resistance of preexisting faults to sliding is largest ("compression" curve in Figure 1). If \(\sigma_1\) is vertical the brittle strength is smallest ("tension" branch in Figure 1). If both \(\sigma_1\) and \(\sigma_3\) are horizontal, the brittle strength may vary between these two extremes [Turcotte and Schubert, 1982; Sibson, 1974].

Ductile Rheology

Recent laboratory data have modified the depth-dependent strength model of Goetze and Evans [1979] only slightly, the main difference being the incorporation of "wet" deformed olivine, i.e., olivine containing trace amounts of water [Chopra and Patterson, 1981; Tsenn and Carter, 1987]. Steady state flow laws for olivine are now well established under both anhydrous conditions and in the presence of water. The creep stress for olivine under conditions that favor dislocation processes may be represented by a powerlaw. Under conditions of low temperatures and high stress the powerlaw breaks down to exponential
behavior [Goetze and Evans, 1979; Tsenn and Carter, 1987]. In
the present study we used a power-law flow law for "wet" olivine
with an average grain size of 1 mm and an exponential law with
data from Tsenn and Carter [1987]. As argued in our discussion
on pore fluid pressures, we do not actually expect the oceanic
lithosphere to be wet but use the present flow law in order to get a
lower bound on the stress magnitude.

Oceanic geotherms are a function of lithospheric age. Temperature
profiles in oceanic lithosphere are calculated using Crough's [1975, 1977]
boundary layer model with a constant basal heat flux and a temperature at the base of the lithosphere
of 1300°C. The basal heat flux is selected on the basis of a fit to
bathymetry and heat flow data [Parsons and Sclater, 1977].
Lithospheric age is estimated from the magnetic anomaly map of
Larson et al. [1985] and the geomagnetic time scale of Harland
et al. [1982].

Focal Mechanisms

Constraints on Brittle Strength From Vertical Slip Components

If a vertical cross section of brittle rock, deforming by plane
strain, is subjected to a horizontal in-plane stress, faults exhibit
normal slip in response to tensile stress and reverse slip due to
compressive stress. As the brittle strength is weaker in tension
than in compression, one could infer whether to use the "tension"
or "compression" branch (Figure 1) from the vertical slip
component. However, if stresses and strains are three-
dimensional, the relation between vertical slip components and
brittle strength is unclear.

The minimum shear stress required for initiation of slip on
preexisting faults is described by Byerlee's [1978] law, whereas
new faults will be created at higher shear stresses described by
Coulomb's [1773] law. Byerlee’s law predicts the most favorable
fault plane orientation as well as the minimum shear stress that is
required for initiation of slip [see Turcotte and Schubert, 1982].
However, if this particular fault is not present, the strength of the
system is higher. Preexisting faults that are less favorably
oriented might become activated at higher stress magnitudes.

We have used a three-dimensional numerical model,
employing Byerlee's and Coulomb's laws, to determine bounds
on the brittle strength from vertical slip components (Appendix
A). We conclude that if the focal mechanism solution indicates a
reverse slip component, the brittle strength is limited by
"compression" and "intermediate" branches in Figure 1. In case
of a normal slip component, the brittle strength ranges from
"intermediate" to "tension". Strike-slip faulting does not provide
any constraint on the brittle strength.

The sign of the vertical slip component is identical on both
nodal planes (see Appendix B). Therefore, putting limits on the
brittle strength by using vertical slip components does not depend
on a proper selection of fault plane and auxiliary plane.

Earthquakes in Ductile Lithosphere

It will become clear in the next section where we explain how
to infer differential stresses from earthquakes, that even if events
occur in ductile rock, stress estimates would be more accurate if
we could limit the range of brittle strengths in the same way we
did for brittle earthquakes. Several models have been proposed
to explain the occurrence of seismic slip at depths where crystal-
plastic processes are thought to occur (we will refer to these
earthquakes as "ductile events"). Sibson [1980] suggests that
brittle frictional sliding within the ductile regime may occur due
to compositional differences. Variation of unstable to stable
frictional slip on deeply penetrating faults is suggested by Tse
and Rice [1986] to explain the occurrence of ductile earthquakes.
Instability caused by injection of fluids into shear zones and fault
gouges, for instance by dehydration reactions, has been put
forward by Raleigh and Paterson [1965] and analyzed by
Shimamoto [1985]. Finally, creep instability, i.e., catastrophic
strain softening due to an increase in strain rate or temperature
[Drowan, 1960; Griggs and Baker, 1968; Hobbs et al., 1986;
Ogawa, 1987], has been put forward to explain ductile
earthquakes. The seismic radiation pattern of most ductile
earthquakes cannot be distinguished from that of a brittle shear
fracture. Currently, it is impossible to select one of the proposed
models on the basis of seismic or rheological data.

In the previous section we discussed how the vertical slip
component of a brittle earthquake can be used to tighten the range of
brittle strengths that are used to estimate upper and lower bounds to the differential stress. The well-established mechanism of
brittle earthquakes was a principal ingredient in this
discussion. For ductile earthquakes the mechanism is unknown,
so that we consider it unwarranted to attribute significance to the
vertical slip component inferred from a double couple
representation of the source. Therefore, if a seismic event occurs
at a depth where rocks are deforming ductily, we take the full
range of brittle strengths into account to estimate the differential
stress ("tension" to "compression" branches in Figure 1).

Differential Stress

The method is designed to yield a measure of stress magnitude
that is directly comparable with results from modeling studies.
Principal stress magnitudes inferred from modeling generally
constitute averages over some (elastic) reference thickness Lref
(in the present study we used Lref=100 km). This reference
thickness has no physical meaning and is only used to facilitate
comparison of stresses derived from force modeling studies with
stresses estimated from earthquakes. The average differential
stress is defined as

\[
\frac{\sigma_1 - \sigma_3}{L} = \frac{1}{L_{ref}} \int_{z=0}^{L_{ref}} (\sigma_1 - \sigma_3) \, dz
\]  

(1)

L is the thermally defined, age dependent thickness of the
lithosphere, i.e., the depth of the 1300°C isotherm in Crough's
[1975; 1977] model. Bending stresses in the interior parts of the
plate are assumed to be negligible.

Method Description

To calculate the average differential stress at an epicentral site,
we note that the strength at the focal depth has been exceeded
locally (see Figure 1). No major strength discontinuities occur
within oceanic lithosphere, so it is a reasonable assumption that
stresses are distributed evenly over the lithospheric thickness.
Weak parts of the lithosphere reach their strength at low stresses,
stronger parts deform elastically until stresses increase sufficiently
to cause deformation either by flow or by fault slip. We conclude
that the strength curve provides an upper limit for differential
stresses at depths where the strength is less than that at the focal
depth. The differential stress in the "elastic core", i.e., the strong
part of the lithosphere where stress is supported elastically, is at
least equal to the stress level at the focal depth [Wortel, 1986;
Wortel and Vlaar, 1988]. After integration of the depth-
dependent differential stress and scaling by Lref we obtain
Fig. 2. Differential stress as a function of age and focal depth, for a strain rate of $10^{-16} \text{s}^{-1}$: (a) for maximum brittle strength in case of reverse slip or strike slip mechanism, (b) for minimum brittle strength in case of normal slip or strike slip mechanism, (c) difference between maximum and minimum stress, a measure for method accuracy, if no constraints on the vertical slip components are available, (d) for maximum brittle strength in case of a normal slip event, and (e) difference between maximum and minimum stress (minimum in Figure 2b). Note that the lower ductile part is not affected by the normal slip constraint, (f) for minimum brittle strength in case of reverse slip, and (g) difference between maximum and minimum stress for focal mechanism with a reverse slip component (maximum differential stress in Figure 2a).
In the following we will refer to the average differential stress as "differential stress".

**Sensitivity of Results to Input Parameters**

Our results are sensitive to two different types of parameters: variables related to the method, that cause uncertainties in the rheology model, and "observational" parameters, like focal depth and lithospheric age. Relevant method-related parameters affecting the depth-dependent rheology model are the pore fluid pressure, the assumed strain rate, and the brittle strength. The effect of decreasing the pore fluid pressure is that the brittle strength and, therefore, our stress estimates increase. As discussed before, we aim at getting a lower bound to differential stresses, within the limits of (what we consider to be) realistic assumptions. We therefore did not include pore pressures smaller than hydrostatic in our calculations.

Figure 1 displays the sensitivity of differential stress to brittle strength and strain rate for a ductile and a brittle event. In Figure 1, hatched areas are a measure of differential stress. The effect of varying the strain rate between $10^{-13}$ s$^{-1}$ and $10^{-12}$ s$^{-1}$ can be significant for ductile earthquakes (Figure 1b). The stress in the elastic core is strongly affected, so that the differential stress is sensitive to a change in strain rate. From Figure 1c it is clear that stresses calculated from events in the brittle part of the lithosphere are relatively insensitive to changes in strain rate.

Both for brittle and for ductile earthquakes the differential stress is very sensitive to the selected brittle strength. In the absence of constraints on the vertical slip component, upper and lower bounds to the differential stress are calculated from "compression" and "tension" strength branches, respectively. Figures 2a and 2b show upper and lower bounds as a function of focal depth and lithospheric age, for a strain rate of $10^{-15}$ s$^{-1}$. A measure of the resolution that can be obtained with our method is the difference between upper and lower bounds to the differential stress (Figure 2c). In older lithosphere we observe differences of several hundreds of megapascals between minimum and maximum differential stresses. Figure 2c can be used to estimate what can be gained by improving the focal depth accuracy after an approximate focal depth range has been determined: an estimated earthquake depth of 20 ± 10 km in 90 Ma lithosphere does not require additional waveform modeling to resolve differential stresses much better. However, if the observed focal depth is 40 ± 10 km, it is worthwhile to try to improve the depth accuracy.

If the focal mechanism has a normal slip component the upper bound to the differential stress is not calculated from the "compression" brittle strength but from the "intermediate" strength branch (Figure 1). The lower bound to the differential stress again follows from using the "tension" brittle strength (Figure 2b). Comparing Figures 2a and 2a, we observe a significant reduction of the maximum differential stress for those focal depths that are in the brittle lithosphere. The exclusion of focal mechanism information for ductile events results in a discontinuity of calculated stresses at the brittle-ductile transition.

The reduction of the maximum differential stress clearly improves the resolution of differential stress estimates for brittle events, as can be seen from comparing Figures 2c and 2d.

In case of reverse slip the "intermediate" in stead of "tension" branch is used to calculate the minimum differential stress. The upper bound to the differential stress follows from the "compression" brittle strength (Figure 2a). Figures 2f and 2g show minimum stress and stress resolution as a function of age and focal depth, assuming that the fault slip has a reverse component. Again, reduction of the range of brittle strengths improves the resolution significantly.

**Differential Stress in the Central Indian Ocean**

The Central Indian Ocean, with its high intraplate seismicity, provides an excellent opportunity to study the relation between earthquakes and the level of intraplate stress. An important motivation to study intraplate stress fields is to improve our understanding of the dynamics of plate motion. With this goal in mind Richardson et al. [1979] performed numerical calculations of the stress field in all major plates. Subsequently, Wortel and Cloetingh and Wortel [1985, 1986].

![Fig. 3. Differential stresses in the Central Indian Ocean at epicentral locations of events listed in Table 1. Thin solid lines from the base of the columns end at epicentral locations. The height from the base of a column is a measure of the differential stress magnitude. Stress magnitudes constitute averages over a reference value (100 km) of the plate thickness. In each column, the darkest grey band corresponds to the inferred differential stress. The thick horizontal line in (or at the top of) the column corresponds with the differential stress calculated by Cloetingh and Wortel [1985, 1986].](image-url)
Cloetingh [1981, 1983, 1985] performed a similar type of numerical modeling, with some important new aspects in their force modeling procedure. In contrast with the earlier models the ridge push was not represented as a line force but as an integrated pressure gradient [Lister, 1975] distributed over all contributing parts of the lithosphere. The slab pull was incorporated as an age-dependent force giving rise to significant lateral variations in the forces representing the subduction process. Whereas the force modeling for the Nazca plate [Wortel and Cloetfngh, 1983, 1985] yielded stress values in agreement with Richardson et al.'s study by Richardson et al. [1979] (see also Richardson [1987, 1989]). This difference stems from a difference in modeling procedure. Richardson et al. [1979] made a variable parameter study of the various plate tectonic forces involved. Cloetingh and Wortel [1985, 1986] treated the driving forces (ridge push and slab pull) as known forces, which could be calculated from kinematic parameters (relative motion along convergent plate boundaries) and the age of the lithosphere involved, and derived the magnitudes of the resistive forces from the equation representing the balance of torques of all forces acting on the plate.

In view of the relevance of such discrepancies for ongoing and future numerical modeling we address this issue by investigating the information which the unusually high seismic activity in the Central Indian Ocean [e.g., Wiens and Stein, 1983, 1984; Bergman and Solomon, 1984, 1985] provides concerning the level of intraplate stress. Previously, only the orientations of the principal stress directions have been used to test the results of stress modeling. For the Central Indian Ocean, good agreement was found between calculated orientations and focal mechanism data [Cloetingh and Wortel, 1986; Bergman, 1986]. Also the orientations of observed long wavelength basement undulations appear to be consistent with the calculated stress orientations [Stein et al., 1989; Petyok and Wiens, 1989].

Large (mL≥4.9) oceanic intraplate earthquakes in the Central Indian Ocean have been the subject of seismological waveform modeling studies by various authors. The earthquakes for which adequate information on depth and focal mechanism were available are listed in Table 1. The focal depth accuracy for shallow teleseismic earthquakes is discussed for various methods by Stein and Wiens [1986]. Depth relocation methods like long-period P waveform modeling [Wiens and Stein, 1983], Rayleigh wave modeling with additional long-period P waveform modeling for shallow events [Wiens and Stein, 1984] and body waveform inversion [Bergman and Solomon, 1984, 1985] all are estimated to have a 2-km accuracy. Pre-WWSSN data, relocated with P, SH, Rayleigh and Love waveform modeling techniques

### TABLE 1. Central Indian Ocean Intraplate Earthquake Data

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth [km]</th>
<th>Vertical Component</th>
<th>Age, Ma</th>
<th>References</th>
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<td>May 25, 1964</td>
<td>-9.1</td>
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<td>8</td>
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<td>24 ± 2</td>
<td>SS</td>
<td>65.5 ± 0.5</td>
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<td>73 ± 9</td>
<td>BS85</td>
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<td>36 ± 11</td>
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a Numbers 7-27 refer to Cloetingh and Wortel [1986], numbers 45-53 have been added to their list.
b Depth from top of crust.
c SS, strike slip; T, thrust; N, normal.
BS85, Bergman and Solomon [1985]; WS83, Wiens and Stein [1983]; SW90, Stein and Weissel [1990]; WS84, Wiens and Stein [1984]; W86, Wiens [1986]; BS84, Bergman and Solomon [1984].
[Wienz, 1986] have 5-km accuracies. Depth phase identifications from short-period $P_{pwP}/P$ arrivals [Stein and Weisell, 1990] are assigned a 5-km accuracy. Vertical slip components are considered significant if the slip direction makes an angle with the strike of more than 10º.

For all these events the differential stress is estimated, first without focal mechanism information taken into account and next with inclusion of such information (see Table 1). The results are displayed in Figure 3 in combination with the numerical model values obtained by Cloetingh and Wortel [1985, 1986]. Figure 3 clearly demonstrates what was already apparent from Figures 2c, 2e and 2g, namely, that hypocentral depth alone does not provide a firm constraint on the level of differential stress. If, however, reliable information on the vertical slip components of an event is available, the large contribution of the brittle slope variation to the uncertainty of the integrated differential stress can be reduced.

Gravity highs attributed to 200-km wavelength basement undulations as a result of lithospheric buckling have been observed in the region between 18ºS and 10ºN and 80ºE and 100ºE. Earthquakes 7, 10, 16, and 18 all occur near highs, so that the effect of flexural bending stresses in this compressive region would be to reduce stresses at the top and increase stress near the bottom of the lithosphere. Differential stress estimates in this region are very high, and the contribution of superimposed flexural bending stresses to the average differential stress generally is not large. More particularly, the stress found from event 7 probably is a lower estimate, and the stress from event 16 is estimated too high. Stresses from the other events are hardly affected by bending stresses, since their epicenters are in the flanks of the buckles. Bathymetric loading by the Chagos-Laccadive Ridge and Ninetyeast Ridge also might cause bending stresses. These would affect the stresses estimated from event 7, 17, 26, and 47-53.

Comparison of Cloetingh and Wortel's [1985, 1986] numerical model values (horizontal bars) with the differential stress ranges in Figure 3 indicates that for events 22, 24, 26, 45, 46, and 47, all near the ridge, the model values are somewhat too high. Only for events 11 and 13 the model values greatly exceed the inferred stress ranges. These events are near the Sumatra-Nicobar-Andaman trench system. All model values for the Central Indian Ocean (with the minor exceptions of one of the events of 10, and 19) fall clearly within the stress ranges. The model values for events 15, 18, and 27 are even near the minima of the seismicity derived stress ranges. We particularly note the good agreement for the high stress values of events 12 and 14.

Since our method yields low-end estimates of the differential stress, our results are more in agreement with those of Cloetingh and Wortel [1985, 1986] than with stress levels found by Richardson [1987, 1989], which are about an order of magnitude lower.

### PACIFIC PLATE STRESSES

Seismicity in the Pacific plate is less abundant than in the Indian plate and focal depth and focal mechanism constraints are relatively poor compared to Indian Ocean data. Table 2 is a selection of $m_0 \geq 4.8$ events we made from World Stress Map data. Figure 4 shows results of differential stress computations from focal depth and focal mechanism.

The highest stresses are estimated for the Samoan-Gilbert-Ralik area from event 6, 7, and 8. Okal et al. [1986] note teleseismically recorded swarms of intraplate seismicity (4.0$\leq m_0 \leq 6.0$) in this region and attribute them to large-scale deformation. However, nearby seamounts probably cause bending stresses so that differential stresses derived from these events are not the intraplate stress field.

Stein [1979] studied event 3 in the northwest of the Pacific and concluded that the epicenter occurred on Emperor Trough, a dead

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth, km</th>
<th>Vertical Component</th>
<th>Age, Ma</th>
<th>Reference</th>
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<tbody>
<tr>
<td>1</td>
<td>March 6, 1965</td>
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<td>-132.8</td>
<td>5 ± 5</td>
<td>N</td>
<td>34 ± 1</td>
<td>O80</td>
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<td>-132.9</td>
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<td>N</td>
<td>34 ± 1</td>
<td>O80</td>
</tr>
<tr>
<td>3</td>
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<td>T</td>
<td>90 ± 10</td>
<td>S</td>
</tr>
<tr>
<td>4</td>
<td>July 29, 1968</td>
<td>-7.5</td>
<td>-143.8</td>
<td>5 ± 5</td>
<td>SS</td>
<td>89 ± 5</td>
<td>O80</td>
</tr>
<tr>
<td>5</td>
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<td>-7.6</td>
<td>-148.1</td>
<td>5 ± 5</td>
<td>N</td>
<td>89 ± 5</td>
<td>O80</td>
</tr>
<tr>
<td>6</td>
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<td>-3.4</td>
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<td>15 ± 3</td>
<td>T</td>
<td>122 ± 3</td>
<td>LO, D83</td>
</tr>
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<td>7</td>
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<td>21 ± 3</td>
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<td>8</td>
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<td>177.5</td>
<td>15 ± 3</td>
<td>T</td>
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</tr>
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<td>9</td>
<td>January 19, 1973</td>
<td>-7.6</td>
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<td>89 ± 5</td>
<td>O80</td>
</tr>
<tr>
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<td>-148.1</td>
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<tr>
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<td>147.1</td>
<td>6 ± 3</td>
<td>T</td>
<td>34 ± 3</td>
<td>BS6, BS</td>
</tr>
<tr>
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<td>13 ± 2</td>
<td>N</td>
<td>18 ± 1</td>
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<td>18 ± 1</td>
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</tr>
<tr>
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<td>-115.9</td>
<td>15 ± 2</td>
<td>N</td>
<td>18 ± 1</td>
<td>WO, D87</td>
</tr>
<tr>
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<td>January 5, 1978</td>
<td>-20.9</td>
<td>-126.9</td>
<td>5 ± 5</td>
<td>T</td>
<td>23 ± 1</td>
<td>O80</td>
</tr>
<tr>
<td>22</td>
<td>July 25, 1978</td>
<td>-20.8</td>
<td>-126.9</td>
<td>5 ± 5</td>
<td>T</td>
<td>23 ± 1</td>
<td>O80</td>
</tr>
</tbody>
</table>

- a Depth from top of crust.
- b SS, strike slip; T, thrust; N, normal.

Table 2. Pacific Intraplate Earthquake Data
spreading center that has also been active as a transform fault between Kula and Pacific plates [Larson and Chase, 1972; Hilde et al., 1976]. Due to the large uncertainties in lithospheric age and focal depth, the difference between maximum and minimum estimated stresses is large. The Caroline plate, north of New Guinea, has been proposed by Weissel and Anderson [1978] to be subducting under the Pacific plate at the Mussau Trench. The epicenter of event 17 lies on the proposed Caroline-Pacific plate boundary and might not be a true intraplate event.

The magnitude of intraplate stresses in the Pacific plate east of 170°W appears to be lower than in the west. The strongest constraint on intraplate stresses in the southeast Pacific comes from event 15. Other earthquakes in this area have less well resolved focal depths. Differential stresses southwest of Baja California also indicate that differential stresses in the eastern Pacific do not exceed the 100 MPa level. Earthquakes in the eastern Pacific that are related to bending stresses due to seamount loads are event 1, 2, and 12 [Okal et al., 1980] and event 11 near Hawaii [e.g., Watts et al., 1985].

DISCUSSION

Strength on Different Time Scales

Jeffreys [1959] observed that high, uncompensated mountain ranges can be supported by the lithosphere and inferred that the strength of the lithosphere must be hundreds of megapascals. Laboratory experiments on both brittle as ductile rocks showed that the lithosphere is indeed capable of bearing stresses of large magnitude on long time scales [Byerlee, 1978; Goetze, 1978]. It was noted by Chinnery [1964] that the stress drop in an earthquake is 2 orders of magnitude lower. Thus, if the stress drop is taken to be a measure of strength, the strength on short time scales (“seismic strength”) seems to be much lower than the strength on geological time scales (“tectonic strength”). The question arises whether we can use long-term strength envelopes to infer stress magnitudes from earthquakes? While further insight into this aspect has to be gained, we adopt the hypothesis concerning the relation between stress, strength (tectology), and earthquake generation proposed by Wortel [1986]. Recognizing that parts of the lithosphere in which the strength is low and where the stress is at or near the strength (comparable with the near-surface part of the oceanic lithosphere) earthquake generation is not observed, he postulated that seismic activity occurs when and where the width (or depth interval) of the anelastically deforming region increases at the expense of the “elastic core”, in other words, when and where the stress reaches the strength for the first time. It was shown that this hypothesis adequately accounts for the distribution of seismic activity in subducting lithosphere [Wortel, 1986; Wortel and Vlaar, 1988]. In this hypothesis the seismic stress drop does not reflect the absolute level of stress but rather a stress adjustment to the equilibrium tectonic stress. Consequently, the seismic strength can be higher than the tectonic strength and earthquakes are envisaged to relax stresses that exceed the long-term strength. The stress drops involved are low, so the long-term strength envelope yields a good approximation of the stresses at the hypocenter.
Therefore, efforts to make the accuracy of focal depths better placed on the stress magnitude. Mechanism is known with confidence, tighter bounds can be placed on the stress magnitude. If the focal uncertainty in differential stresses is large. If the focal mechanism is known with confidence, if the faulting is pure strike-slip, or if an earthquake occurs in ductile lithosphere, the uncertainty in differential stresses is large. If the focal mechanism is known with confidence, tighter bounds can be placed on the stress magnitude.

Numerical calculations of the stress field in the Central Indian Ocean by Richardson et al. [1979] and Cloetingh and Wortel [1985, 1986] yield order of magnitude differences. From our analysis we find that seismicity data require a stress level comparable with that calculated by Cloetingh and Wortel [1985, 1986]. Stress levels about an order of magnitude lower, as advocated by Richardson [1987, 1989] are less in agreement with differential stresses found in this study.

Appendix A

We investigate the relation between the vertical slip component of an earthquake and the brittle strength. Obviously, both are related by the stress field; slip occurs in the direction of maximum resolved shear stress if the strength is exceeded.

In general, the brittle strength according to a linear friction law can be written in terms of the difference between maximum and minimum principal stresses

\[ \sigma_1 - \sigma_3 = \frac{2 S_0}{\mu} + \frac{2 \mu (q - \lambda)}{\mu + \mu_0} \lambda \]

\[ q = \frac{\sigma_1 + \sigma_3}{2 \sigma_0} \cdot \frac{\sqrt{1 + \mu_0^2 + \lambda \mu}}{\mu + \mu_0} \leq q \leq \frac{\sqrt{1 + \mu_0^2 - \lambda \mu}}{\mu + \mu_0} \]

cohesion \( S_0 \), coefficient of friction \( \mu \), overburden load \( \sigma_0 \), and pore fluid coefficient \( \lambda \). We assume that one of the principal stress directions is vertical and equal to the overburden pressure. Upper and lower limits to the brittle strength and \( q \) correspond to in-plane compression and in-plane tension, respectively, or more precisely, to vertical \( \sigma_1 \) and \( \sigma_3 \). Therefore, if we would know the principal stress directions at the epicenter, we could select a single brittle strength curve to estimate differential stress. Unfortunately, focal mechanism solutions only give principal deformation quadrants, not principal stress directions.

Fig. A1. Vertical fault slip component as a function of horizontal stresses \( \sigma_x \) and \( \sigma_y \). The grey shaded areas indicate stresses at which slip along preexisting faults can take place. At lower stresses, in the internal part of the figure, the deformation is purely elastic. The failure criterion puts an upper bound to the stress level that can be reached. In the light grey shaded area the predominant vertical component is reverse slip. In the dark grey area normal slip is observed. The shape of the figure is invariant to depth; the figure simply shrinks or expands at smaller or greater depths (numbers on the axes correspond to a depth of 13 km).
To determine the relation between focal mechanism and applied stress, we performed a number of synthetic tests. For various three-dimensional stress fields we calculated the orientations of preexisting fault planes on which the conditions for slip are met; that is, the threshold value of shear stress has been reached or exceeded according to Byerlee's law. Horizontal stresses are limited by the fault strength of brittle rock, which was calculated by Coulomb's [1776] law.

Figure A1 shows a typical result of our modeling. Three basic response types are indicated; if horizontal tectonic stresses (\(\sigma_x\) and \(\sigma_y\)) are small deformation is elastic. If stresses increase slip on preexisting faults (if present) may occur. If stresses cannot be accommodated on preexisting faults, new ones may be created at even higher stresses. Faulting types typical for particular stresses may be recognized [Anderson, 1951; Sibson, 1974]; if both horizontal stresses are compressive mainly reverse faults become activated. Normal faulting typically occurs in response to horizontal tensile stresses. The switch of slip with normal component to slip with reverse component occurs approximately midway the transitional domain where \(\sigma_z\) is vertical.

From our synthetic experiments we conclude that if a focal mechanism has a significant reverse component, the brittle strength is bounded by "compression" and "intermediate" curves in Figure 1. If a normal slip component has been observed, the brittle strength lies between "tension" and "intermediate" curves. A mechanism with a pure strike-slip character does not give any constraints on the brittle strength, so that the full strength range has to be taken into account.

**APPENDIX B**

Given the orientations of unit length \(T\), \(P\), and \(B\) axes we will show that the vertical slip component on both nodal planes has the same sign. Let \(X\) be north, \(Y\) east, and \(Z\) vertically downward. To specify the orientation of a particular vector, we use angles \(\phi\) and \(\delta\). \(\phi\) is the angle between the projection of the vector on the horizontal \(X-Y\) plane and the \(X\) axis, measured clockwise. \(\delta\) is the clockwise angle between the horizontal projection of the vector and the vector itself. \(P\), \(T\), and \(B\) axes are typically given in lower hemisphere stereographic coordinates, so that \(\delta \in [0, \pi]\) and \(\phi \in [0, 2\pi]\).

We define a matrix \(R\) that rotates \(X\) to the \(B\) axis \((\phi_B, \delta_B)\):

\[
R = \begin{bmatrix}
\cos\phi_B \cos\delta_B & -\sin\phi_B \cos\delta_B & \sin\delta_B \\
\sin\phi_B \cos\delta_B & \cos\phi_B \cos\delta_B & -\sin\phi_B \\
\sin\delta_B & \cos\delta_B & 0
\end{bmatrix}
\]

If we let \(R^{-1}\) work on \(T\) axis \((\phi_T, \delta_T)\) and \(P\) axis \((\phi_P, \delta_P)\), both \(P' = R^{-1} P\) and \(T' = R^{-1} T\) will lie in the \(Y-Z\) plane. We now define a third angle \(\gamma\), to specify the angle in the \(Y-Z\) plane between the \(P'\) or \(T'\) axis and \(Z\). We consider two cases: \(T'\) closest to the vertical, i.e., \(\gamma \in [-\pi/4, \pi/4]\), and \(P'\) closest to the \(Z\) axis.

If \(T'\) is closest to the system in the north direction \(B\) axis, we define a second rotation matrix

\[
V = \begin{bmatrix}
1 & 0 & 0 \\
0 & \cos\gamma \gamma & -\sin\gamma \gamma \\
0 & \sin\gamma \gamma & \cos\gamma \gamma
\end{bmatrix}
\]

so that \(T'' = R V T'\). \(T''\) is the vertical unit vector. Slip vectors in the system of vertical \(T''\) axis and north direction \(B'\) axis are \(S_{1}'' = (0, 1/2 \sqrt{2}, -1/2 \sqrt{2})\) and \(S_{2}'' = (0, -1/2 \sqrt{2}, 1/2 \sqrt{2})\). Slip vectors in the original system of \(P\), \(B\), and \(T\) axes can be calculated from

\[
S_{1} = R V S_{1}'' \\
S_{2} = R V S_{2}''
\]

Evaluation of the vertical slip components in the original system of \(P\), \(B\), and \(T\) axes yields

\[
S_{1} = -\sqrt{2} \cos \gamma (\cos \gamma + \sin \gamma) \\
S_{2} = -\sqrt{2} \cos \gamma (\cos \gamma - \sin \gamma)
\]

for \(\gamma \in [-\pi/4, \pi/4]\). For the given range of \(\gamma\), both vertical slip components have the same sign.

By the same approach it can be shown that if the \(P'\) axes is closest to the vertical in the system of north directing \(B\) axis we have

\[
S_{1} = -\sqrt{2} \cos \gamma (\cos \gamma - \sin \gamma) \\
S_{2} = -\sqrt{2} \cos \gamma (\cos \gamma + \sin \gamma)
\]

for \(\gamma \in [-\pi/4, \pi/4]\), so that both vertical slip components have either dip-slip or thrust components.

**Acknowledgments.** During this research, R.G. was financially supported by AWON, the Earth Science branch of the Netherlands Organization for Scientific Research (NWO). M.J.R.W. and S.A.P.L.C. received partial support from NATO grant 0148/87.

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